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Seismic Study of an Oceanic Ridge Earthquake Swarm in the Gulf of California

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Summary

Detailed seismic investigation of an unusually intense earthquake swarm which occurred in the northern Gulf of California during March 1969 has provided new information about seismic processes which occur on actively spreading oceanic ridges and has placed some constraints on the elastic wave velocities beneath them. Activity during this swarm was similar to that of a foreshock–mainshock–aftershock sequence, but with a 'mainshock' composed of over 70 events with magnitudes between 4 and 5·5 occurring in a 6-hr period about a day after swarm activity was initiated. 'Aftershocks', including many events greater than magnitude 5, continued for over two weeks. Near-source travel-time data indicate all sources located are within 5–10 km of each other and that hypocentres are confined to the upper crust. Teleseismic *P*-delays for rays travelling beneath this ridge may be interpreted in terms of an upper mantle with compressional velocities 5–10 per cent less than normal mantle to a depth of 200 km. Average apparent stresses for all swarm events studied are very similar, show no consistent pattern as a function of time, and are close to values obtained from other ridges. The focal mechanism solution shows a large component of normal faulting. An apparent non-orthogonality of nodal planes common to this mechanism solution and to normal faulting events on other ridges disappears when the indicated low upper mantle velocities beneath the source are taken into account.

A survey of recent seismicity (post 1962) in the northern Gulf suggests seismic coupling across about 200 km between adjacent inferred spreading ridge segments.

Surface waves from these Gulf Swarm earthquakes have amplitudes from one to two orders of magnitude greater than Northern Baja California events with similar short period body wave excitation.

1. Introduction

Seismic investigations have provided considerable stimulus and support to the hypotheses of sea-floor spreading and rigid plate tectonics. The focal mechanisms of earthquakes on mid-ocean ridges determined by Sykes (1967, 1968), Tobin & Sykes (1968), Banghar & Sykes (1969) have been a striking confirmation of Wilson's (1965) transform faulting hypothesis and have given strong stimulus to the development of

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the concepts of plate tectonics by Morgan (1968), McKenzie & Parker (1967) and Le Pichon (1968).

Other seismic investigation has outlined several important characteristics of mid-ocean ridges. Surface and body wave travel-time studies of the Mid-Atlantic ridge near Iceland by Trygvasson (1962, 1964) suggest a zone of anomalously low seismic velocity extends from the Mohorovicic discontinuity to a depth of the order of 200 km. Talwani *et al.* (1965) found that gravity and seismic refraction data from profiles over the Mid-Atlantic ridge were consistent with a low density region whose upper surface is at the crust–mantle interface beneath the ridge crest but deepens away from the axial zone. Recently Wyss (1970b) has applied the method of comparing the excitation of short period body waves and long period surface waves to infer that relatively low average stresses are acting in the source region of ridge earthquakes. Sykes (1970), in a study of the available seismic data pertinent to spreading ridges has recently pointed out the close relationship between ridges, earthquake swarms, normal faulting and vulcanism.

The Gulf of California is one of the few places on the globe where sea-floor spreading and lithospheric generation is occurring in a region accessible to close-in land-based seismic observation. The Gulf of California has been spreading at a half-rate of 3 cm yr^{-1} for the past 4·5 My, as demonstrated by Larson, Menard & Smith (1968) in their examination of the magnetic lineations at the mouth of the Gulf. That it is actively spreading at present is clearly demonstrated by the seismicity and focal mechanism investigations of Sykes (1968). The crustal structure varies from oceanic at the mouth of the Gulf to continental shelf type (crustal thickness 20–25 km) near the north end, but with considerable lateral variation indicated normal to its axis (Phillips 1964; Thatcher & Brune 1969).

This paper is a detailed study of an unusually intense earthquake swarm which occurred in the northern Gulf near 31° 10′N, 114° 26′W during March 1969, and includes analysis of seismic data recorded less than 60 km from the epicentres of this oceanic ridge swarm. In addition the seismicity of the Gulf since 1962 is surveyed and its relationship to swarm activity is examined.

2. Swarm activity and gulf seismicity

The northern third of the Gulf of California is shown in the bathymetric map of Fig. 1, modified from Fisher, Rusnak & Shepard (1964). The closed basins striking roughly north-east are presumed spreading centres, representing a median depression at the ridge axis, and the elongate north-west striking bathymetric lines which connect them mark the traces of transform faults. Two focal mechanisms obtained by Sykes (1968) and shown on the map are consistent with this interpretation of the bathymetry and the known geology. The three seismograph stations used in the location of swarm events are Rio Hardy (RHM), a permanent station about 130 km to the north of the epicentres, and San Felipe (SF-) and El Golfo (EG-), two portable seismographs each located about 50 km from the swarm. The locations determined are within about 5 km of the dot shown on the map and lie between Wagner Basin, a closed bathymetric depression on the sea floor, and Consag Rock, an andesitic knob protruding above the surface of the Gulf about 20 km to the west of the basin.

Swarm activity began at 08 16 GMT on 20 March with an event of local magnitude 5·2 (all magnitudes quoted for the March 1969 swarm are local magnitudes as determined at Pasadena unless otherwise noted). At 08 17 and 08 23 shocks of magnitude 5·7 and 5·2 respectively occurred, and during the next 18 hr about a dozen events greater than magnitude 4 were detected. On 21 March at 02 34 very intense activity was initiated and in the following 6-hr period 75 quakes with $M_L \geqslant 4$ occurred, 20 of these being greater than magnitude 5. Activity fell to half these values in the following

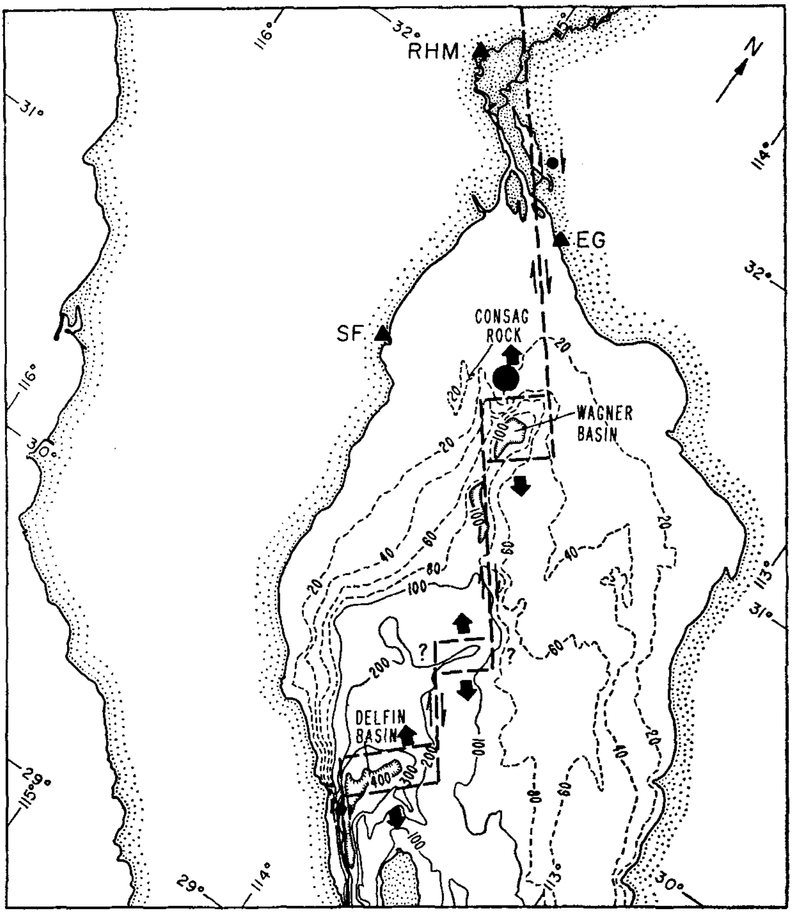

FIG. 1. Bathymetric map of the northern Gulf of California, modified from Fisher *et al.* (1964). Closed basins striking roughly north-east are presumed spreading centres and elongate north-west trending bathymetric lines are inferred transform fault segments. Local seismograph stations and best swarm location are shown, along with two focal mechanisms by Sykes (1968).

6-hr, but activity continued for over a week, with over 200 events greater than magnitude 4 and 40 above magnitude 5 recorded at Pasadena by 28 March.

The seismic activity of earthquake swarms and aftershock sequences may be conveniently displayed as a function of time by a plot of cumulative seismic moment versus time. In the dislocation theory of seismic sources (Burridge & Knopoff 1964; Aki 1966), seismic moment is directly proportional to the fault area times the average slip across the fault surface. It is a more precise seismic source parameter than either magnitude or energy, and may be determined with factors of 2 to 3 accuracy from measurements of the amplitudes of long period surface waves (Aki 1966). It has been extensively used by Brune (1968), Wyss & Brune (1968), and Wyss (1970a) in seismic source studies in the western United States. Fig. 2 is a plot of cumulative seismic moment versus time for the Gulf swarm. For the purposes of this plot, the approximate empirical relation between local magnitude and seismic moment of Wyss & Brune (1968) for the western United States suffices to relate magnitude to moment for all Gulf events with $M_L \geqslant 4{\cdot}0$. The accuracy of this equation for the northern Gulf region is verified by the values determined for the

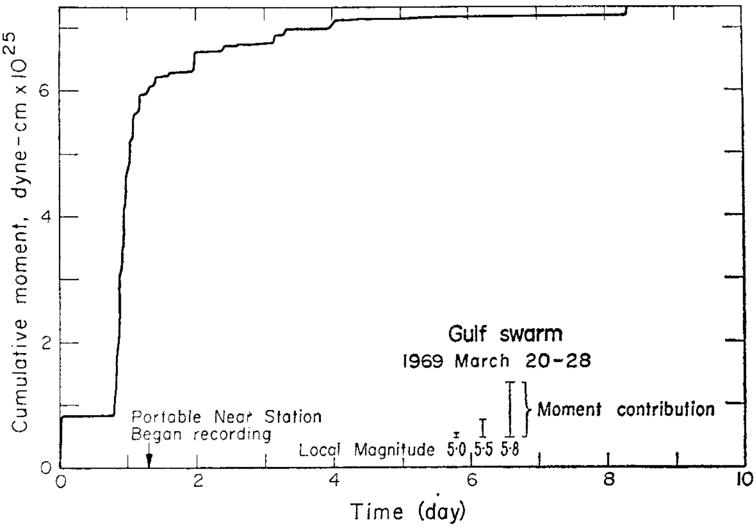

FIG. 2. Cumulative seismic moment plotted as a function of time for main Gulf swarm—Local earthquake magnitude has been translated into moment using the magnitude-moment relation of Wyss & Brune (1968). Note that seismic moment is proportional to displacement across a dislocation (fault surface) times fault area.

moments of swarm events using the amplitude spectra of long period Rayleigh waves, discussed below (Section 6). To be noted in Fig. 2 is that the peak 6-hr period of activity a day or so after the initiation of the swarm accounts for over three-quarters of the cumulative moment of the entire sequence–rather like a foreshock–mainshock–aftershock pattern, but with the ' mainshock ' made up of an extraordinary series of magnitude 4 and 5 earthquakes occurring during one 6-hr period. The cumulative moment during this period corresponds about to a single magnitude 6·2 event.

The frequency–magnitude relationship has been used by many authors to characterize regional seismicity, and in fact is a measure of the mean earthquake magnitude. The slope, or b-value, of such a plot for the March 1969 Gulf swarm is about 0·90 using Pasadena local magnitudes (208 samples). Using the smaller USCGS sample for the swarm (78 events), their body wave magnitudes, the b-value is about 1·35. The difference in the two b-values may be reflecting the differences in the two magnitude scales as well as the differences in the number of samples. Evernden (1970) observed that b-values from m_b data were consistently larger than M_L b-values for world-wide earthquakes but only for those over magnitude 5-$\frac{1}{2}$ or so. The b-value of 0·90 is close to values found for the San Andreas system in southern California by Allen *et al.* (1965) using Pasadena M_L data.

The seismicity of the Gulf of California from 1 January 1962 to date has been surveyed using USCGS epicentre listings. It should be noted that detection of Gulf events is strongly affected by seismograph station distribution, and hence relatively more events of magnitude less than $m_b = 5·0$ will be reported in the northern Gulf as compared to the southern Gulf (where there are few established stations close to the active region). Even in the northern Gulf it is clear from study of the March 1969 swarm that many events with $m_b \leqslant 4·5$ or so are missed in CGS listings. Still, with these limitations and with only an 8 year sampling of northern Gulf seismicity, an interesting pattern of activity is evident. Table 1 is a listing of all sequences of more than 5 events occurring during a time interval of a month or less since 1962 in the northern Gulf. Both the Wagner basin and the Delfin basin area, about 150 km to the south (see Fig. 1) are regions of recurring activity, and furthermore the seismicity

Table 1

Summary of Northern Gulf Seismic Activity since 1962 *January* 1 *from USCGS epicentre listings*

	Time interval	Approximate location	Number of events listed	Body wave magnitudes (m_b)
1	18–23 Nov 1963	30 °N 113·5 °W Delfin Basin	14	4·1—5·7*
2	3–4 Feb 1964	31·3 °N 114·3 °W Wagner Basin	9	4·0—4·8
3	27 Aug–18 Sept 1967	31·2 °N 114·2 °W Wagner Basin	7	4·0—4·4
4	5–6 Dec 1967	30·5 °N 114·2 °W Northern Delfin Basin Region	9	3·8—5·0
5	2–18 Feb 1969	30 °N 113·5 °W Delfin Basin	11	4·4—4·8
6	20 Mar–6 Apr 1969	31·3 °N 114·2 °W Wagner Basin	78	3·9—5·5 (14 events with $m_b \geqslant 5$)

*First events of this sequence is the $m_b = 5 \cdot 7$ shock for which Sykes (1968) determined a strike slip mechanism.

in the two areas appears to be coupled: the earthquake sequences are paired, and when activiy is initiated in one basin it is followed in one to three months by activity in the other. No significant activity occurred in the area between the two basins in the time interval between the two sequences. Five of the six sequences are swarm-like, each with many events of roughly similar magnitude rather than one clearly dominant event preceded and/or followed by much smaller shocks. The exception, sequence 1, had a main shock of $m_b = 5 \cdot 7$ with a strike-slip focal mechanism (Sykes 1968) followed by aftershocks up to magnitude $m_b = 5 \cdot 3$. The choice of sequences of five events or more was arbitrary but convenient—it would be hazardous to draw any conclusions with fewer events than this because of the biased omission of smaller magnitude shocks (three ‘sequences’ of four events each were detected, and two of these *could* have been interpreted as preceded or followed by two to three events in the other basin).

The location of sequence 1 very close to the bathymetric depression of the Delfin basin indicates a small segment of transform fault in the vicinity and demonstrates that even when tectonic and thermal conditions approach closely those of an actively spreading ridge, the earthquake mechanism and the aftershock sequence are still characteristic of a ‘normal’ mainshock–aftershock series along a transform fault. This argues against attributing ridge swarms to inhomogeneities in physical properties (Mogi 1963) unless these inhomogeneities are very localized. The close association of ridge swarms with magmatic activity and the creation of new lithosphere may provide the localized stress concentrations necessary to cause swarms (Mogi 1963; Sykes 1970), but the details of swarm mechanism are far from being resolved.

The Imperial Valley, in southernmost California approximately 200 km north of the Gulf, experiences considerable seismic activity including swarms (Richter 1958), and a search of USCGS and Pasadena local bulletins was made for sequences such as those found in the Gulf. The findings are summarized in Table 2. It should be kept in mind that the largest of these events recorded by the USCGS had a body wave magnitude of only 4·6, and that many shocks recorded in the Imperial Valley by the Caltech network would not have been reported had they occurred in the Gulf, where only USCGS listings are used in this comparison of seismicity. The events in

Table 2

Summary of Imperial Valley Seismicity from 1962 *January* 1 *Pasadena Local Bulletin and USCGS listings*

	Time interval	Approximate location	Number of events listed	Magnitudes (M_L)
1	27 Oct–2 Nov 1963	33° 15′—115° 40′	19 (9 by CGS)	3·0—4·4
2	16–17 June 1965	33° 05′—115° 40′	18 (12 by CGS)	3·0—4·4
3	17–22 Dec 1968	33° 02′—115° 50′	10 (Mainshock only by CGS)	2·5—4·7 (M_L = 4·7 and aftershocks)
4	31 July–6 Aug 1969	32° 55′—115° 33′	13 (None reported by CGS)	1·8—3·3

Table 2 locate close to Obsidian Buttes, Quaternary volcanic rocks which protrude through the thick sediments of the valley near the south end of the Salton Sea. A comparison of Tables 1 and 2 demonstrates that between the Valley and the Gulf differences exist in both the frequency of shocks and in their maximum magnitudes, at least for the time interval examined here.

Also, besides swarm events, large strike-slip earthquakes, along presumed transform faults, e.g. the 1940 El Centro earthquake (M_L = 7·1) and the 1966 shock in the Colorado delta (M_L = 6·3), characterize the seismicity of the Imperial Valley–Colorado delta region. It is *possible* to correlate two of the valley sequences with the northern Gulf pairs, but the two others are not simply correlated. If any seismic coupling does exist between the northern Gulf and the Imperial Valley, it is not demonstrated by the data examined here.

To the south of Delfin basin, Gulf seismicity (USCGS) reveals no correlative patterns with northern Gulf activity. A sequence in the mid-Gulf region (approx. 27 °N, 110 °W) from 29 June to 6 July 1964 consisted of 13 earthquakes including an m_b = 6·0 event on 5 July preceded and followed by several shocks of up to magnitude 5·4. The earthquakes were located 60 km from the nearest inferred spreading centre, the focal mechanism determined by Sykes (1968) was strike-slip, and the sequence had a foreshock–mainshock–aftershock character. A recent sequence further south (17–19 August 1969 at approx, 25 °S, 109 °W) began with two large events (m_b = 5·7 and 6·1) within two minutes of each other and had 18 smaller aftershocks. Its location and character suggest transform faulting. These two sequences and the ones listed in Table 1 were the only ones detected in the Gulf of California since 1962. Though intermediate cases are not definitely precluded, available evidence from Gulf earthquakes suggest that it is possible to make a distinction between oceanic ridge and transform fault sequences on the basis of location, maximum magnitude, time-magnitude behaviour, and focal mechanism.

3. Location of March 1969 swarm events

Two portable seismographs located about 50 km from the epicentres of these events were set up about a day and a half after the swarm began. Fig. 3 shows their location and in Fig. 2 the time of their installation with respect to swarm activity is noted. The region of the swarm is illustrated in Fig. 3, along with the locus of possible epicentres and their corresponding hypocentral depths determined using only *S–P* times at the two closest stations. The crustal structure used is from a seismic refraction

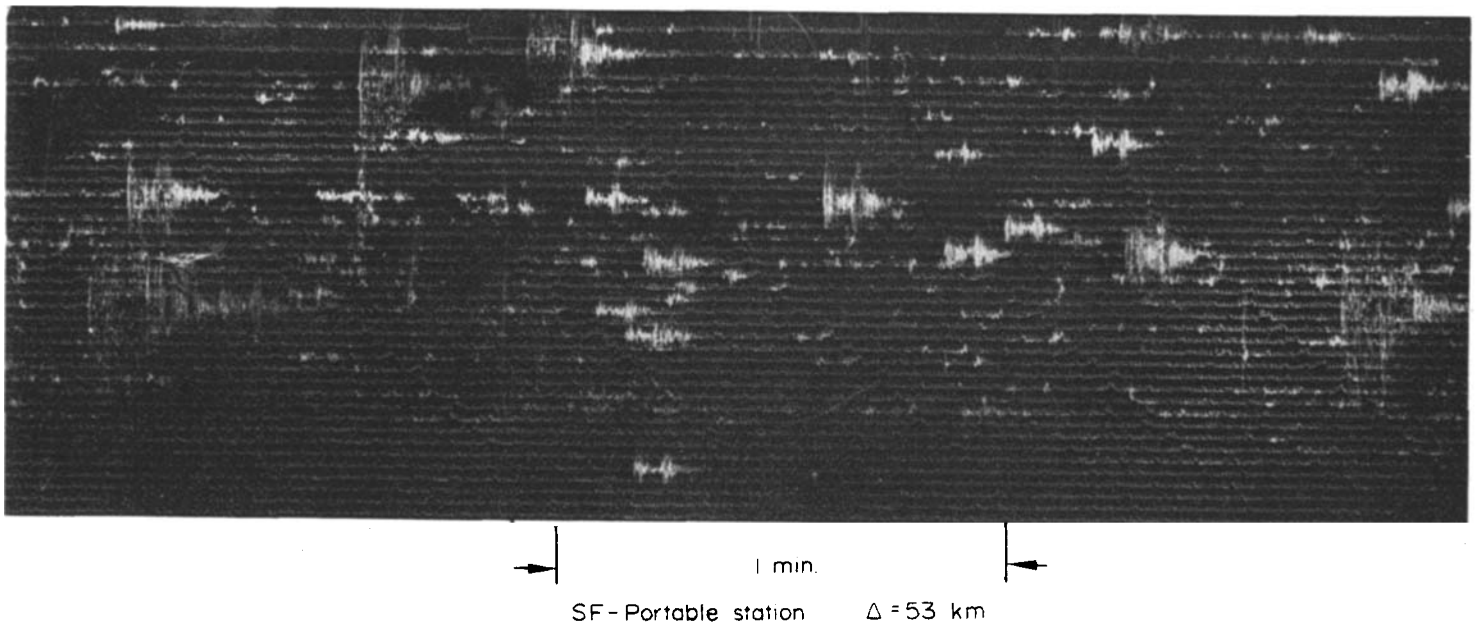

FIG. 4. Sample seismic record from portable station SF-, located about 50 km from the events shown (note S–P times of about 7 s). Absolute timing accuracy is about $\pm 0 \cdot 1$ s.

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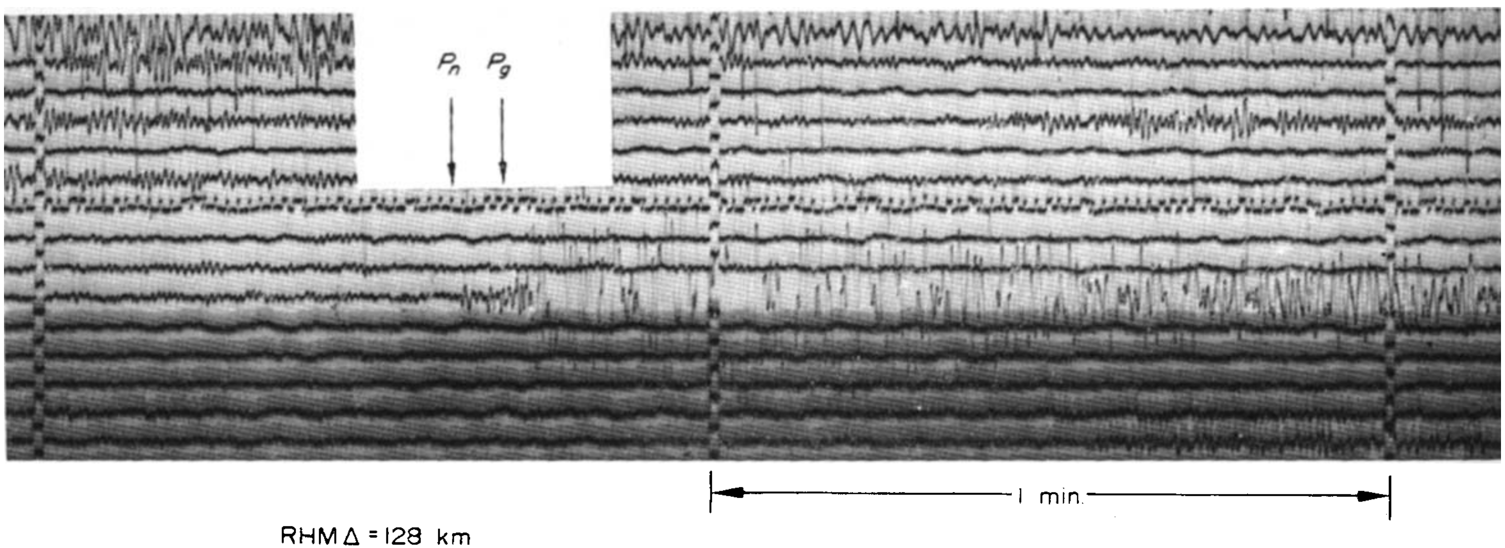

FIG. 5. Typical swarm event recorded at RHM, permanent station, 130 km to the north of epicentres. Seismograph is short-period Benioff. Note the first arrival P_n, and the strong second arrival about 5 s later, interpreted here as the crustal phase P_g. These data provide the strongest body-wave constraint on the focal depth of swarm events.

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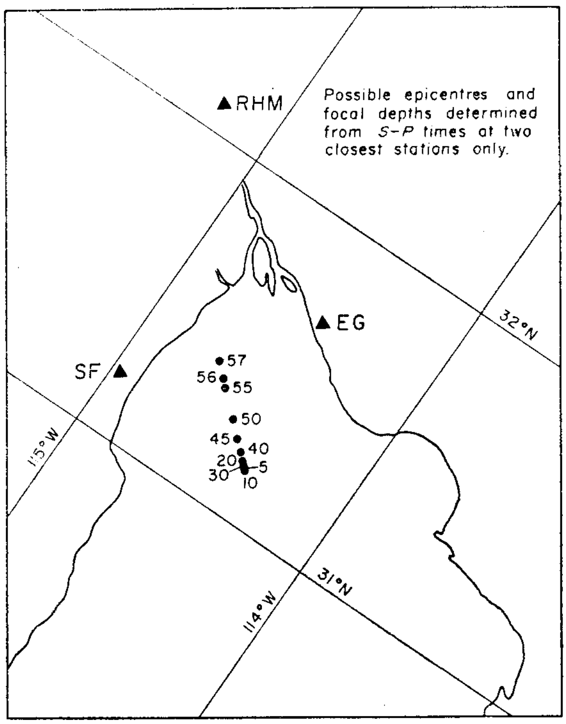

FIG. 3. Possible epicentres (with hypocentral depths which correspond to them) obtained using S–P times at two nearest stations only.

survey in the region by Phillips (1964), and Poisson's ratio was taken to be 0·25 in order to determine P-travel times from S–P intervals. Fig. 4, which is a sample seismic record from portable station SF-, shows many events with S–P times within a few tenths of 7·0 s. Other records from the portable stations show similar consistency. From readings of these two stations alone, it is seen that focal depth is not well determined, and any depth less than about 55 km will satisfy the data. However there are several other important observations which help constrain the depth.

The focal depth is most strongly constrained by readings at RHM, approximately 130 km to the north-west of the swarm, as well as by readings at more distant (200–300 km) Caltech stations in southern California. Fig. 5 is a short period record of a typical swarm event recorded at RHM. The critically refracted P_n arrival is followed about 5 s later by a strong second arrival, interpreted here as the crustal phase P_g. Using Phillips' (1964) crustal structure (crustal thickness 25 km, P_n velocity 7·8 km s^{-1}.) travel-time curves were constructed for a range of possible focal depths down to 50 km. No fit to the data was possible for a source in the upper mantle. The best fit to the near-source body wave observations was provided by placing the sources in the upper crust at a depth of 7 km—the P_g–P_n times at RHM and in southern California could only be satisfied if the hypocentres were located above the 6·70 km s^{-1} layer in Phillips' crustal model, and 7 km satisfied the data best. The travel-time curves and the data to which they were fit are shown in Fig. 7 along with the velocity model used in constructing the plot. Note that over 60 pieces of data support the interpretation made here. In addition, readings of P_g–P_n at Caltech stations in the Imperial Valley region (Glamis $\Delta = 225$ km; Hayfield $\Delta = 290$ km) fit this interpretation well. Moderate changes in crustal structure would change the results only slightly and we feel that a hypocentre in the upper crust for swarm events

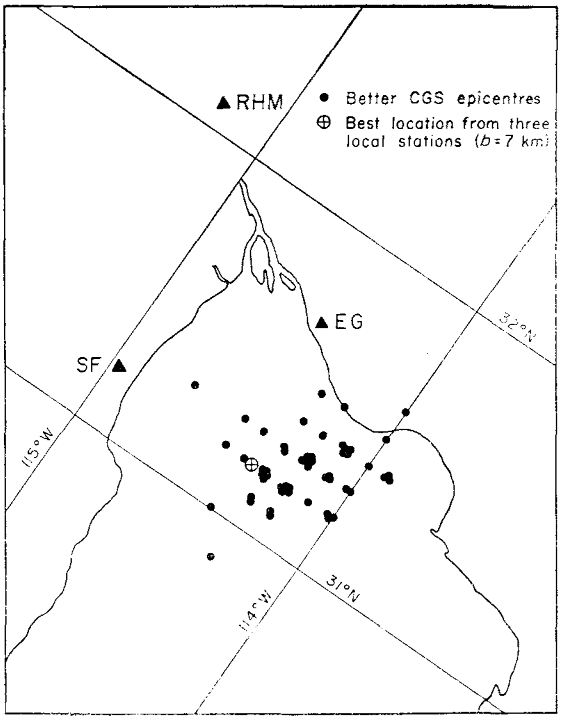

FIG. 6. Region of swarm of March 1969, showing better located USCGS epicentres (shown without asterisk in their listings and determined by them to the nearest tenth of a degree). Also shown are the local seismograph stations used in locating swarm events with better precision, as well as the best location found for many events using these near-source travel times.

is well established. A seismic study is presently being carried out in the Imperial Valley to clarify the relationship between the San Andreas system and the ridge-transform fault pattern of the Gulf of California, which may extend into mainland North America as far as the Imperial Valley–Salton Sea area. Fig. 6 is a map of the same area as Fig. 3 showing the better determined epicentres of the USCGS (their locations are given to the nearest tenth of a degree), as well as the best location determined in this study using many travel-time observations from the three local stations. The broad distribution in USCGS epicentres is not believed to be real—the consistently similar S–P times recorded at local stations indicate all events studied here occur within approximately 5 km of each other. Our preferred location lies within 10–15 km of the centre of the cluster of USCGS epicentres. If the true focal depths were greater than 45 km, the USCGS epicentres would have to be consistently in error by more than 30 km, greater even than mislocation errors in island arc regions. From what is known of the relative importance of lateral inhomogeneities in mid-ocean ridges and in island arcs, such a large mislocation appears unlikely—a relatively hot, low-velocity upper mantle beneath a spreading ridge would tend to produce travel-time delays which have little preferred azimuth distribution at teleseismic distances, and in such events a trade-off between origin time and focal depth will occur, so that locations may be too deep, but there should be little epicentral bias.

Another line of evidence which can place limits on the focal depths of earthquakes is the shape of the seismic surface wave spectrum. The dependence of focal depth upon spectral amplitude is most marked for surface waves of Rayleigh type, and is a consequence of the variation in excitation as a function of depth. For each wave period of the fundamental mode, the horizontal excitation has a node (zero crossing) at a depth of about one-quarter wavelength, while the curve of vertical excitation versus depth has a low gradient at the quarter wavelength depth. Beyond a half wavelength both horizontal and vertical excitation decrease rapidly with increase in depth. Thus the cumulative effect of focal depth alone on Rayleigh wave spectra is to produce:

(1) a minimum at a wave period (T_m) which corresponds to the focal depth (Z_0), i.e. $T_m \approx [4Z_0/C(T_m)]$, where $C(T_m)$ is the phase velocity at period T_m, and

(2) a fall-off in the spectrum for periods less than about $[2Z_0/C(T_m)]$.

For focal depths less than 60 km or so these diagnostic changes in spectral shape occur for periods between 30 and 10 s, and are less pronounced for normal faulting than for strike-slip mechanisms. Differences in focal mechanism can alter the depth effects by up to a factor of 2 or so (Tsai 1969) and for some mechanism parameters the spectral minimum may be absent at certain azimuths from the source (Harkrider, personal communication). In addition, the spectrum in this period range is subject to several other complications, including the effect of source dimension, uncertainty of Q-correction, and contamination by short period higher modes with group

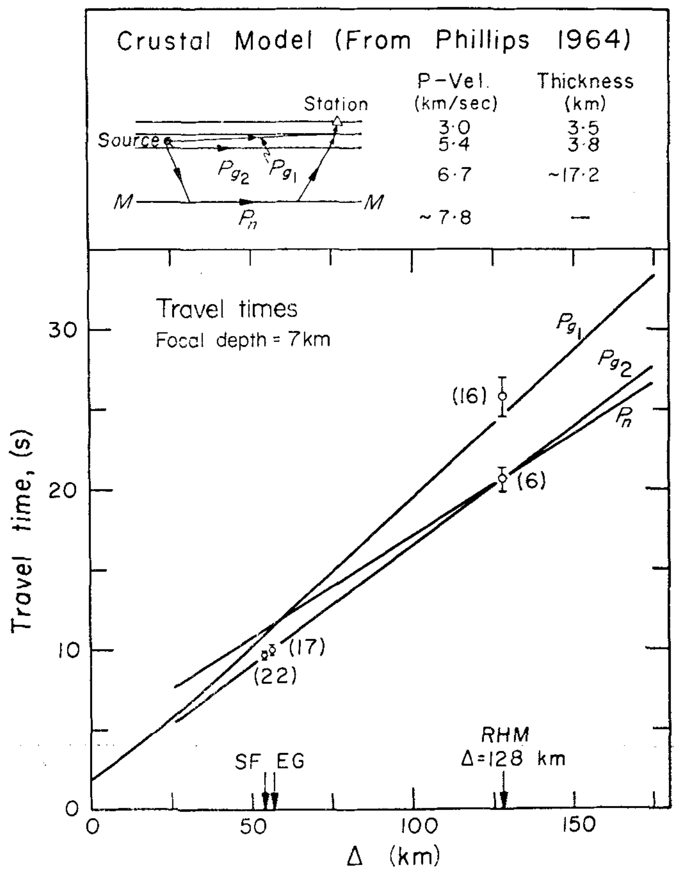

FIG. 7. Travel-time curve which provides a good fit to all the near station data. Crustal model is from Phillips (1964). The data are shown with their error bars (2 standard deviations) and the number in parentheses refers to number of pieces of independent data which each point represents.

velocities similar to those of the fundamental mode. For the Gulf swarm earthquakes, location of many events within a region of less than 10 km or so indicates finiteness should not seriously effect the spectra. Similarly, contamination by higher Rayleigh modes in the critical period range should not occur for continental Rayleigh waves, though oceanic Rayleigh dispersion studies by Sykes & Oliver (1964) and Kovach & Anderson (1964) suggest that such contamination would occur for oceanic paths. The fundamental mode group velocity curve for oceanic Rayleigh waves is not well defined for periods less than 20 s, and higher modes are frequently excited in the period range 10–30 s with group velocities which make them difficult to separate from fundamental mode wavetrains. The uncertainty in Q-correction at short periods complicates the interpretation as well, but may be minimized by checking spectral shape at short distances, where corrections are smallest. Tsai (1969) shows several convincing cases for determining focal depth of continental strike-slip sources using continental Rayleigh wave spectra, but his results are more equivocal for oceanic paths and sources, and we believe this is a result of the difficulties mentioned above.

The Rayleigh wave spectra from a number of Gulf events recorded at several locations and azimuths in the United States have been measured. The event shown in Fig. 8 is typical of many others, and its shape suggests the source is shallow. The spectrum is corrected for instrument response but not for Q, since even within its uncertainties attenuation will not significantly alter the shape of the spectrum shown here, since the epicentral distance is only about 8°. Geometrical spreading and dispersion corrections (independent or period) have not been made for the data in the plot illustrated.

Spectral amplitudes for periods less than about 10 s are less certain than for longer periods because of uncertainties in Q, but this uncertainty has been minimized

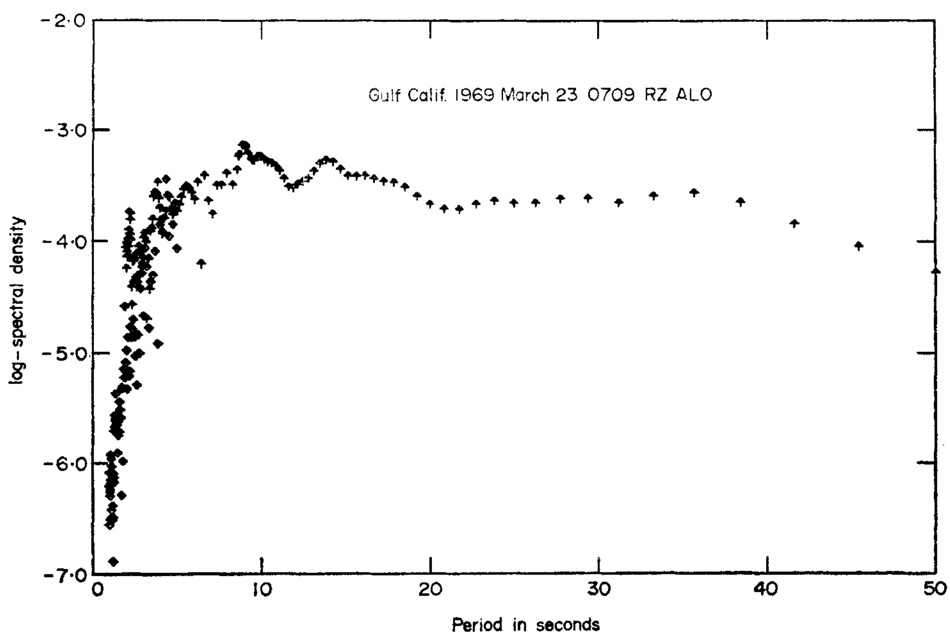

FIG. 8. Rayleigh wave spectrum from a typical swarm event (1969 March 23 07 09, $M_L = 4 \cdot 6$) recorded at WWSSN station ALQ (Albuquerque, New Mexico, $\Delta = 7 \cdot 63°$). Long and short period records have been patched together to include all of the fundamental mode spectrum between 1- and 50-s period. Short period higher modes in the period range 5–10 s have been excluded from seismograms analysed on the basis of their expected group velocities. Spectrum corrected for instrument response but not for geometrical spreading or attenuation.

because of the short path from the Gulf to ALQ (New Mexico). This significance of the short period amplitudes is that their relatively high excitation (even without attenuation corrections, which would increase their values) indicate a source in the shallow crustal layers. A deep crustal or upper mantle source would not significantly excite these short periods.

4. Origin times and *P*-wave residuals

For a number of larger 1969 swarm events recorded at the portable stations, it was possible to compare the origin times determined from near-source readings with those determined by the USCGS using teleseismic *P*-times. Table 3 shows the results of this comparison for nine events listed by the CGS for 21 March 1969, and the average difference in origin time is about three seconds. This difference could be the results of *P*-delays beneath the source and/or incorrect assignment of focal depth by the USCGS. In the absence of near-source travel-time readings the trade-off between origin time and focal depth makes it difficult to accurately determine origin time and hence detect low seismic velocities beneath the source. However, all but one of the events listed by the CGS were constrained to 33 km depth, and the one shock with an assigned focal depth of 2 km had an origin time which was *early* by about one second compared to the locally determined one. Thus the discrepancies in both focal depth and origin time are resolved, given that the sources are located in the upper crust, as found above. With origin times precisely determined from local stations, it is possible to more accurately examine teleseismic travel-time data for evidence of low seismic velocity beneath this spreading ridge segment. Arrival times for *P*-waves from Gulf swarm events reported in USCGS EDR bulletins were used with the near-source origin times to obtain travel times which were compared with those of several of the more recently determined *P*-tables (Carder, Gordon & Jordan 1966; Herrin *et al.* 1968; Johnson 1969). Only five stations at epicentral distances beyond 30° consistently reported times for better located events listed in Table 2, but residuals from these stations, shown in Table 4, demonstrate that *P*-delays of about two seconds occur in this distance range. The range in residual at an individual station reflects the slightly differing travel-time tables of the different investigators. *P*-wave

Table 3

Comparison of USCGS Origin Times with those determined from S–P times at portable stations 1969 *March* 21

M_b	USCGS			This paper			Observed −CGS
4·6	15	07	13·6	15	02	11·0	−2·6
5·1	15	57	42·0	15	57	43·4	+1·4[1]
4·7	16	29	40·4	16	29	34·2	−6·2
4·7	17	55	47·2	17	55	42·7	−4·5
5·2	18	00	20·6	18	00	17·2	−3·4
4·3	20	11	51·0*	20	11	49·4	−1·6*
4·2	20	36	50·3*	20	36	46·4	−3·9*
4·2	21	05	38·1*	21	05	33·9	−4·2*
4·6	23	03	25·2*	23	03	22·2	−3·0*
						Average =	−3·1 s (for better determined events)

[1]USCGS focal depth is 2 km for this event
* Less well-determined epicentres

Table 4

P-wave travel-time residuals for swarm events of 1969 *March* 21

Station	Δ (degrees)	Observed-*P* table+Station correction (seconds)
COL	39·54	+2·4→3·2
MBC	45·19	+2·0→2·8
SJG	45·40	+1·5→2·3
ALE	54·53	+0·5→1·2
NUR	82·68	+2·3→3·1
	Average =	+1·7→2·5 (±0·6) s

Residuals are referred to *P*-times of Carder *et al.*, Herrin *et al.*, and Johnson (assumed depth = 7 km)
Station corrections from Cleary & Hales (1966)

station corrections tabulated by Cleary & Hales (1966) add between 0·4 and 0·7 s to the *P*-delays for the stations listed in Table 2. If the travel time delay is confined to the upper 200 km of the mantle, then *P*-wave velocities beneath this ridge are 5–10 per cent less than those for a normal mantle.

5. Focal mechanism

The poor azimuth distribution of WWNSS stations with respect to the Gulf of California, as well as the relatively low body wave magnitude of even the largest swarm shocks (m_b = 5·5 was the largest) made it difficult to accurately determine *P*-wave focal mechanisms for this swarm. In addition, many of the largest events occurred during the peak period of activity and *P*-wave first motions were obscured by preceding events. However, the first large event of the swarm (1969 March 21 08 17 41·9, m_b = 5·4, M_L = 5·7) was registered clearly at enough stations to approximately define nodal planes, and Fig. 9 is a projection of its *P*-first motions on the

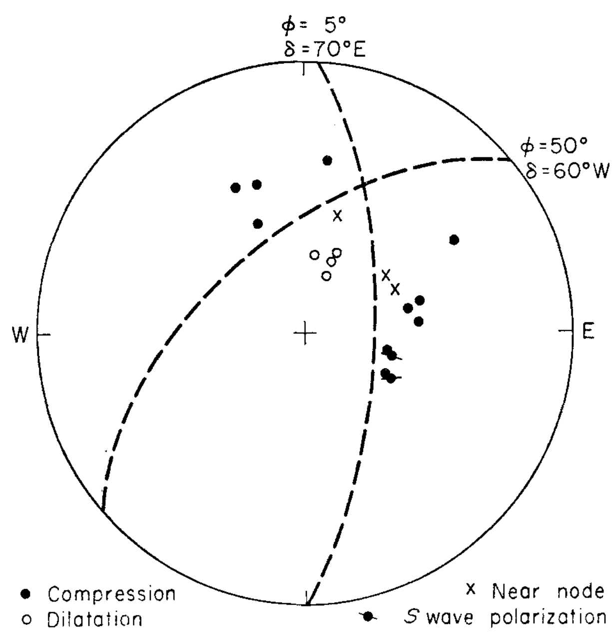

FIG. 9. Focal mechanism for Gulf swarm event on 1969 March 20 08 17 41·9. Projection of the lower hemisphere of focal a sphere constructed from Jeffreys–Bullen velocity structure.

lower hemisphere of the focal sphere. S-wave polarization data supplement the P observations. Though the nodal planes are not *precisely* defined, they are determined within confined limits, and note that two orthogonal nodal planes cannot be fitted to the observations with this projection. Furthermore, a survey of published ocean ridge normal faulting mechanisms (by the Lamont group) with nodal planes confined by the data reveals that this is generally the case, and the derived nodal planes cross at 60 to 70 degrees rather than at right angles. The nodal planes for strike-slip mechanisms in oceans are almost invariably orthogonal. The discrepancy may be simply explained by the fact that the projection most commonly employed and the one used here utilizes extended distance tables constructed from the Jeffreys–Bullen velocity structure, which differs considerably from that which is presumed to occur directly beneath ridges. With the P-velocity decrease that is suggested by the travel-time residuals discussed in the previous section, simple refraction at the Mohorovicic discontinuity and/or the top of the low velocity layer can account for the shifting of each nodal plane by 10–15 degrees. Note that this refraction will not effect strike-slip mechanisms because of their symmetry.

6. Seismic moments and apparent stresses

The seismic moment and average apparent stresses were determined for 15 of the March 1969 swarm events. Moments were obtained using measurements of the amplitude spectral densities of long period Rayleigh waves recorded on WWSSN seismographs at several stations in the central and eastern United States (ALQ, FLO, GEO, OXF, WES) suitably corrected for instrument response, and the effects of propagation and radiation pattern (Ben Menahem & Harkrider 1964). Seismic energy (E_s) was defined from the Gutenberg and Richter energy–magnitude relationship, and the average apparent stress ($\eta\bar{\sigma}$) determined from the formula (Aki 1966)

$$\eta\bar{\sigma} = \frac{\mu E_s}{M_0},$$

where M_0 is the seismic moment, μ is the shear modulus (3×10^{11} cgs) and η the seismic efficiency factor. Table 5 contains the results. Events are arranged chronologically, but no clear pattern emerges between the apparent stresses of earlier and

Table 5

Seismic moment and apparent stresses for March 1969 swarm events

Date	Time	Moment, M_0 10^{24} dyne-cm	M_L	E_s 10^{20} dyne-cm	Apparent stress ($\eta\bar{\sigma}$) bars
21 March	1010	15·25	5·5	1·12	2·2
	1224	0·70	5·3	0·56	24·0
	1557	1·00	5·1	0·29	8·7
	1629	2·04	5·0	0·20	3·0
	1800	1·15	5·2	0·41	10·7
	2303	0·75	4·9	0·14	5·6
22 March	0725	3·56	5·5	1·12	9·2
	1823	3·72	5·2	0·41	16·4
	2256	0·75	4·8	0·10	13·7
23 March	0709	0·22	4·6	0·05	6·7
	1132	0·22	5·2	0·41	20·4
	1539	0·60	5·2	0·41	16·4
24 March	0902	0·88	5·3	0·56	15·6
28 March	1519	1·20	5·3	0·56	14·0
3 April	1055	1·00	4·9	0·14	4·2

later events. Small events in the months before and after the swarm show no significant differences in surface wave excitation, and thus there is no indication of any systematic differences in stresses acting in the source region before and after the swarm. For the 15 events studied, the average of the apparent stresses is 11 ± 5 bars, in good accord with the average for ridges and transform faults in the Atlantic and north-east Pacific determined by Wyss (1970b), his average for 27 events being 17 ± 12 bars.

7. A comparison with earthquakes in Northern Baja California

Northern Baja California, about 100–200 km west of the Gulf swarm earthquakes, is a region of high seismic activity in a tectonic environment very different from that of the Gulf of California. Northern Baja California is underlain by granitic batholitic rocks of regional extent as compared to the oceanic or transitional nature of the crust beneath the Gulf. The relatively low excitation of surface waves from earthquakes in this region was first noted by Brune, Expinoza & Oliver (1963) and was interpreted by Wyss & Brune (1968) as possibly due to relatively high tectonic stresses acting in the source region of these shocks. Just how profound the difference can be is illustrated in Fig. 10, which compares Rayleigh waves from a Gulf swarm shock and a north Baja event of comparable magnitude recorded with identical instrument magnification at Oxford, Mississippi. The average apparent stress for this event and for others in north Baja is about 200 bars, and small source dimensions are also indicated. A detailed study of the source parameters of these earthquakes is presently underway and will be reported in a later paper.

Though not yet investigated in detail, the importance of presumed high stress earthquakes in the problem of discriminating earthquakes from underground nuclear explosions is well known. For a given magnitude, both show relatively low excitation of long period surface waves. Studies of the M_s–m_b discrimination criterion do not

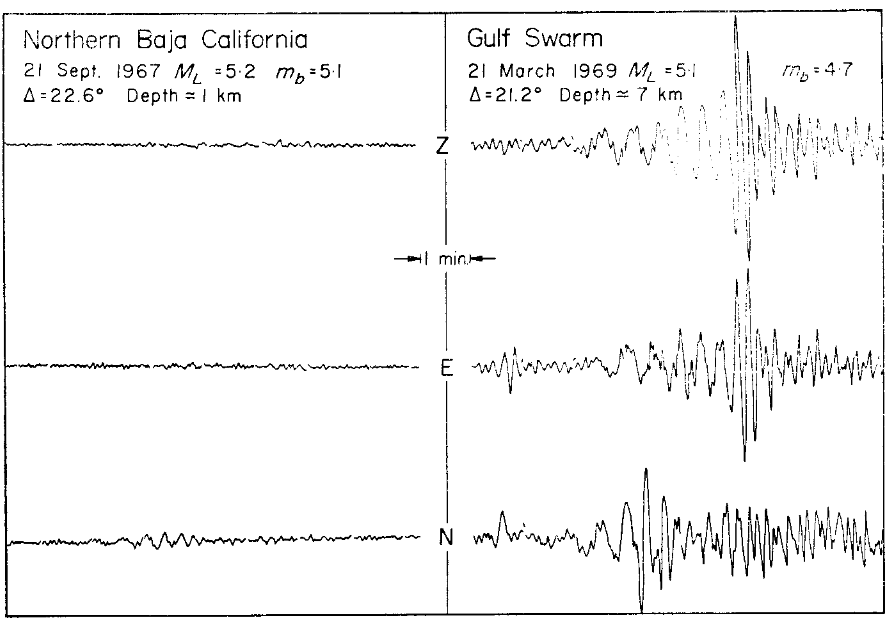

FIG. 10. Comparison of long period Rayleigh wavetrains recorded at OXF (Oxford, Mississippi) for a Gulf swarm shock and an event in northern Baja California of about the same local magnitude.

consider significant numbers of high stress earthquakes. Basham (1969) *was* able to discriminate the north Baja shock shown in Fig. 10 from explosions at the Nevada Test Site using an M_s–m_b plot. However, a large population of high stress earthquakes could seriously affect such a discrimination tool: the difference between high stress earthquakes and possible nuclear explosions from the same region would be quite subtle, and existing discrimination methods may not in general be capable of resolving them.

8. Conclusions

Seismic study of a remarkable earthquake swarm in the Northern Gulf of California has provided some details on seismic activity and structure beneath a spreading oceanic rise. Swarm sequences here are characterized by shallow hypocentral depths, predominantly normal faulting, and a distinctive activity as a function of time (viz. Fig. 2). Teleseismic *P*-delays from these swarm sources suggest anomalously low upper mantle velocities beneath this spreading ridge, and compressional velocities 5–10 per cent less than normal mantle to a depth of 200 km are consistent with the observations.

It is also of some interest to compare the seismic characteristics of spreading ridge earthquakes with those in adjacent parts of the Northern Gulf region. Within the Gulf itself, recent seismicity listings suggest seismic coupling across about 200 km between two adjacent inferred spreading ridge segments in the Northern Gulf. Data examined here demonstrates no such coupling between the Northern Gulf spreading centres and centres of Quaternary volcanism in the Imperial Valley region. Finally, it was observed that earthquakes in Northern Baja California typically have surface wave amplitudes one to two orders of magnitude less than Gulf earthquakes with similar short period excitation.

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References

Aki, K., 1966. Generation and propagation of G waves from the Niigata earthquake of June 16, 1964, 2, Estimation of earthquake moment, released energy, and stress-strain drop from G wave spectrum, *Bull. Earthquake Res. Inst., Tokyo Univ.*, **44,** 73.

Allen, C. R., St Amand, P., Richter, C. F. & Nordquist, J. M., 1965. Relationship between seismicity and geologic structure in the southern California region, *Bull. seism. Soc. Am.*, **55,** 753.

Banghar, A. & Sykes, L. R., 1969. Focal mechanism of earthquakes in the Indian Ocean and adjacent areas, *J. geophys. Res.*, **73,** 632.

Basham, P. W., 1969. Canadian magnitudes of earthquakes and nuclear explisions in the southwestern United States, *Geophys. J. R. astr. Soc.*, **17,** 1.

Ben Menahem, A. & Harkrider, D. G., 1964. Radiation patterns of seismic surface waves, *J. geophys. Res.*, **69,** 2605.

Brune, J. N., Espinoza, A. & Oliver, J., 1963. Relative excitation of surface waves by earthquakes and nuclear explosions in the California–Nevada region, *J. geophys. Res.*, **68,** 3501.

Brune, J. N., 1968. Seismic moment, seismicity and rate of slip along major fault zones, *J. geophys. Res.*, **73,** 777.

Burridge, R. & Knopoff, L., 1964. Body force equivalents for seismic dislocations, *Bull. seism. Soc. Am.*, **54,** 1874.

Carder, D. S., Gordon, D. W. & Jordan, J. N., 1966. Analysis of surface foci travel times, *Bull. seism. Soc. Am.*, **56,** 815.

Cleary, J. & Hales, A., 1966. An analysis of the travel times of *P* waves to North American stations, in the distance range 32° to 100°, *Bull. seism. Soc. Am.*, **56,** 467.

Evernden, J. F., 1970. Study of regional seismicity and associated problems, *Bull. seism. Soc. Am.*, **60,** 393.

Fisher, R. L., Rusnak, G. A. & Shepard, F. P., 1964. Submarine topography of the Gulf of California, chart accompanying *Marine Geology of the Gulf of California*, T. H. Van Andel and G. G. Short, Eds, Amer. Assoc. Petrol. Geologists, Tulsa, Oklahoma.

Herrin, E. (Ed.), Arnold, E. P., Bolt, B. A., Clawson, G. E., Engdahl, E. R., Gordon, D. W., Hales, A. L., Lobdell, J. L., Nuttli, O., Romney, C., Taggart, J. & Tucker, W., 1968. 1968 seismological tables for *P* phases, *Bull. seism. Soc. Am.*, **58,** 1193.

Johnson, L. R., 1969. Array measurements of *P* velocities in the lower mantle, *Bull. seism. Soc. Am.*, **59,** 973.

Kovach, R. L. & Anderson, D. L., 1964. Higher mode surface waves and their bearing on the structure of the earth's mantle, *Bull. seism. Soc. Am.*, **54,** 161.

Larson, R. L., Menard, H. W. & Smith, S. M., 1969. The Gulf of California: A result of ocean-floor spreading and transform faulting, *Science*, **161,** 781.

Le Pichon, X., 1968. Sea-floor spreading and continental drift, *J. geophys. Res.*, **73,** 3661.

McKenzie, D. P. & Parker, R. L., 1967. The North Pacific: an example of tectonics on a sphere, *Nature*, **216,** 1276.

Mogi, K., 1963. Some discussions on aftershocks, foreshocks, and earthquake swarms—the fracture of a semi-infinite body caused by an inner stress origin and its relation to earthquake phenomena, 3, *Bull. Earthquake Res. Inst., Tokyo Univ.*, **41,** 615.

Morgan, W. J., 1968. Rises, trenches, great faults, and crustal blocks, *J. geophys. Res.*, **73,** 1959.

Phillips, R. P., 1964. Seismic refraction studies in Gulf of California, in *Marine Geology of the Gulf of California*, T. H. Van Andel and G. G. Shor, Eds, Amer. Assoc. Petrol. Geologists, Tulsa, Oklahoma.

Richter, C. F., 1958. *Elementary Seismology*, W. H. Freeman and Company, San Francisco.

Sykes, L. R., 1967. Mechanism of earthquakes and the nature of faulting on the mid-ocean ridges, *J. geophys. Res.*, **72,** 2131.

Sykes, L. R., 1968. Seismological evidence for transform faults, sea-floor spreading, and continental drift, in *Proceedings of NASA Symposium, History of the Earth's Crust*, R. A. Phinney, Ed., Princeton University Press, New Jersey.

Sykes, L. R., 1970. Earthquake swarms and sea-floor spreading, *J. geophys. Res.*, **75,** 6598.

Sykes & Oliver, 1964. The propagation of short period seismic surface waves across oceanic areas, 2, Analysis of seismograms, *Bull. seism. Soc. Am.*, **54,** 1373.

Talwani, M., Le Pichon, X. & Ewing, M., 1965. Crustal structure of the mid-ocean ridges, 2, computed model from gravity and seismic refraction data, *J. geophys. Res.*, **70,** 341.

Thatcher, W. & Brune, J. N., 1969. Surface waves and crustal structure in the Gulf of California, Baja California, and Sonora, *Trans. Am. geophys. Union, EOS*, **50,** 240.

Tobin, D. & Sykes, L. R., 1968. Seismicity and tectonics of the north-east Pacific Ocean, *J. geophys. Res.*, **73,** 3821.

Trygvasson, E., 1962. Crustal structure of the Iceland region from the dispersion of surface waves, *Bull. seism. Soc. Am.*, **52,** 359.

Trygvasson, E., 1964. Arrival times of *P* waves and upper mantle structure, *Bull. seism. Soc. Am.*, **54,** 727.

Tsai, Y. B., 1969. Determination of focal depths of earthquakes in the mid-oceanic ridges from the amplitude spectra of surface waves, Ph.D. thesis, Massachusetts Institute of Technology, Cambridge.

Wilson, J. T., 1965. A new class of faults and their bearing on continental drift, *Nature*, **207,** 343.

Wyss, M. & Brune, J. N., 1968. Seismic moment, stress and source dimensions for earthquakes in the California–Nevada region, *J. geophys. Res.*, **73,** 4681.

Wyss, M., 1970a. Observation and interpretation of tectonic strain release mechanism, Ph.D. Thesis, California Institute of Technology, Pasadena.

Wyss, M., 1970b. Apparent stresses of earthquakes on ridges compared to apparent stress of earthquakes in trenches, *Geophys. J. R. astr. Soc.*, **19,** 479.